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Meridional carbon dioxide transport in the northern North Atlantic

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Abstract

Combination of estimated water transport and accurate measurements of total carbon dioxide (TCO_2) on a hydrographic section at 58°N allows the assessment of meridional inorganic carbon transport in the northern North Atlantic Ocean. The transport has been decomposed into contributions from the large-scale baroclinic overturning, the Ekman transport, baroclinic and a barotropic eddy terms, and an estimated contribution of the East Greenland Current. These terms are $-0.27 \cdot 10^6$, $+0.03 \cdot 10^6$, $+0.03 \cdot 10^6$, $+0.10 \cdot 10^6$, and $+0.05 \cdot 10^6$ mol s^{-1}, respectively, which result in a total southward inorganic carbon transport of only $-0.06 \cdot 10^6$ mol s^{-1}. An order of magnitude estimate of the meridional transport of dissolved organic carbon (DOC) has shown that in general this term cannot be ignored in the total carbon flux, this being $+0.04 \cdot 10^6$ to $+0.16 \cdot 10^6$ mol s^{-1} at 58°N. A simple carbon budget has been formulated for the temperate North Atlantic, using our flux estimates as well as those of Brewer et al. (1989). This budget shows that the divergence of the meridional carbon flux, connected with the freshwater balance of the ocean may be of the same order of magnitude as the divergence of the total inorganic carbon flux. For an accurate estimate of the total carbon budget of the ocean it will be necessary to take both the DOC transport and the effects of the freshwater balance into account.

Keywords: meridional transport; TCO_2; North Atlantic

1. Introduction

The global climate system is considerably influenced by the North Atlantic Ocean. Warm surface waters are driven by the rotation of the wind stress, resulting in a net northward transport to the Nordic Seas. There heat loss to the atmosphere causes cooling and subsequent descent of water into the deep North Atlantic as Iceland–Scotland Overflow Water (ISOW) and Denmark Strait Overflow Water (DSOW), driven by density gradients. These water types form the main ingredients of the North Atlantic deep-water complex which flows southwards through the Atlantic Ocean. These two compensating flows form the basis of the oceanic "conveyor belt" concept (Broecker and Peng, 1982). The large-scale overturning in this conveyor belt transports heat and fresh water in meridional direction and has also been suggested to be a major pathway by which anthropogenic carbon dioxide, taken up from the atmosphere by the surface ocean, is transported into the deep ocean.

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Since the onset of the industrial revolution, mankind has exponentially increased the amount of fossil fuel CO_2 released into the atmosphere, currently $\sim 14.3 \cdot 10^6$ mol s^{-1} (5.5 Gt a^{-1} C; Anderson et al., 1991). Estimates of the capacity of the surface ocean to take up this CO_2 vary between $3.9 \cdot 10^6$ and $5.2 \cdot 10^6$ mol s^{-1} (1.5 and 2.0 Gt a^{-1} C; Tans et al., 1990; Sarmiento, 1991). These estimates are based on global circulation models and identify the need for terrestrial uptake of CO_2 to balance the global carbon budget. Later this terrestrial sink has been shown to be less important (Siegenthaler and Sarmiento, 1993; Sundquist, 1993) than previously postulated by Tans et al. (1990), thus further complicating the problem of assessing the global carbon budget.

In order to quantify the role of the oceans in the carbon cycle an accurate understanding is needed, not only of the net exchange between ocean surface and the atmosphere, but also of accurate estimates of the amounts of carbon transported by the large-scale circulation in the ocean basins. Estimates of the meridional total carbon dioxide (TCO_2) flux exist for the subtropical North Atlantic (Brewer et al., 1989), and for the tropical Atlantic (Broecker and Peng, 1992; Keeling and Peng, 1995). Martel and Wunsch (1993) tried to assess the meridional TCO_2 transport in the whole North Atlantic, but failed to obtain statistically significant results from their inversion of a finite-difference model based on a non-synoptic set of hydrographic sections.

Brewer et al. (1989) investigated the meridional TCO_2 transport in the Atlantic Ocean across a section at 25°N and through Florida Straits. A total of nine stations had been sampled for TCO_2, oxygen, alkalinity, and nitrate. Data had been interpolated and extrapolated on density surfaces to cover the whole section used in their calculations. Using the published transport rates of Hall and Bryden (1982) and Roemmich and Wunsch (1985), which were based on inverse modelling with a zero net meridional flow constraint, they arrived at a net southward transport of $-0.68 \cdot 10^6$ mol s^{-1} (-0.26 Gt a^{-1} C), presumably being mainly pre-industrial CO_2, since man-made radiochemical tracers in the deep waters still had not reached this latitude. Such pre-industrial net southward transport in the deep Atlantic Ocean is thought to balance atmospheric transport from the southern to the northern hemisphere, i.e. the pre-industrial atmospheric CO_2 in the southern hemisphere would have been higher than in the northern hemisphere. Based on this result the North Atlantic was deemed to play a minor role in the uptake of atmospheric CO_2. However, Tans et al. (1990) found the temperate North Atlantic Ocean between 15°N and 50°N to take up $0.78 \cdot 10^6$ mol s^{-1} (0.30 Gt a^{-1} C) of atmospheric CO_2, a combination of pre-industrial and anthropogenic CO_2, with a subsequent transport into the deep ocean.

Broecker and Peng (1992) corrected their inferred carbon transport for the effect of fossil fuel addition and organic material breakdown. By assuming no net meridional flow and a thermohaline overturning of 20 Sv (1 Sv = 10^6 m^3 s^{-1}) and a typical difference between surface water and deep water of 80 μmol kg^{-1} they came to a southward inter-hemispheric transport rate in the tropical Atlantic Ocean of $-1.6 \cdot 10^6$ mol s^{-1} (-0.6 Gt a^{-1} C). With similar arguments, but an assumed thermohaline overturning of only 13 Sv and a more detailed hydrography Keeling and Peng (1995) arrived at a southward inter-hemispheric carbon transport rate in the tropical Atlantic of $-1.05 \cdot 10^6 \pm 0.48 \cdot 10^6$ mol s^{-1} (-0.40 ± 0.18 Gt a^{-1} C). These pre-industrial transport estimates are already larger than the net uptake by the North Atlantic Ocean estimated by Tans et al. (1990) who did not correct for fossil fuel addition. Balancing the global carbon budget would thus still need the presence of a terrestrial sink though probably with a smaller capacity for uptake. However, Stoll et al. (1996-this issue) arrived at a southward meridional transport at 58°N in the Atlantic Ocean of $-0.16 \cdot 10^6$ mol s^{-1} (-0.06 Gt a^{-1} C) from a similar budget estimate of the overturning in the Oceanic Conveyor belt in the northern North Atlantic. With the value of $-0.68 \cdot 10^6$ mol s^{-1} (-0.26 Gt a^{-1} C) of Brewer et al. (1989) this suggests a net oceanic uptake from the atmosphere of only $0.52 \cdot 10^6$ mol s^{-1} (0.20 Gt a^{-1} C) in the temperate North Atlantic, 30% less than the above mentioned value of Tans et al. (1990).

Martel and Wunsch (1993) tried to invert a finite-difference circulation model of the whole North Atlantic instead of a single section in order to estimate the ocean circulation and to derive the meridional heat, O_2, and TCO_2 fluxes. The model used nonsynoptic data from hydrography, current meters, sub-surface floats and wind stress, interpolated on a 1° × 1° grid. However, the resulting meridional TCO_2 flux turned out to be not statistically significant. Martel and Wunsch (1993) ascribed this to the scarcity of data, the smoothing required by the use of nonsynoptic multiple cruise data, and the insufficient resolution of the model grid.

The massive fixation of inorganic carbon by marine phytoplankton is an important factor in the uptake of CO_2 by the surface ocean. Particulate organic matter (POM) as well as dissolved organic carbon (DOC) derived from phytoplankton result in an additional pool of carbon, albeit organic, which is also transported by the ocean circulation. A comparison of the order of magnitude of advective fluxes of organic and inorganic carbon will give more insight into the relative importance of the organic and inorganic carbon pools for the carbon budgets.

In this paper we will present an estimate of the meridional transport of TCO_2 in the northern North Atlantic at ~ 58°N. We will also discuss the relative importance of the transport of DOC for the meridional carbon transport, and how to use such transport estimates in the construction of a regional carbon budget for the North Atlantic Ocean. We will hereby discuss the effects of a non-zero meridional water mass flow, which is generally not considered in carbon flux estimates, and its relation with the freshwater budget of the ocean.

During the DUTCH-WARP project RV "Tyro" carried out a hydrographic survey of WOCE Hydrographic Programme section AR7E from Ireland to Greenland in April 1991 as well as a number of high-resolution sections in the Iceland Basin in late April to early May (Fig. 1). The AR7E section at ~ 58°N stretched from the Irish continental shelf to the edge of the East Greenland Current, which could not be surveyed due to heavy pack ice. The nominal station distance along AR7E was 30 nautical miles (55.6 km), reduced to 15 nautical miles (27.8 km) over steep topography. Water from all 24 Niskin bottles in the rosette sampler was collected for the determination of nutrients, O_2, and TCO_2 for all stations in section AR7E. The resulting vertical resolution was 200 m at mid-depth, diminishing to 50 m in the near-bottom and near-surface layers. These hydrographic and biogeochemical data from section AR7E are a unique data set for estimating advective water mass transport rates and associated meridional chemical fluxes in the northern North Atlantic. The distribution of TCO_2 along section AR7E is shown in Fig. 2a. An extensive hydrographic description of all sections is given by Van Aken and De Boer

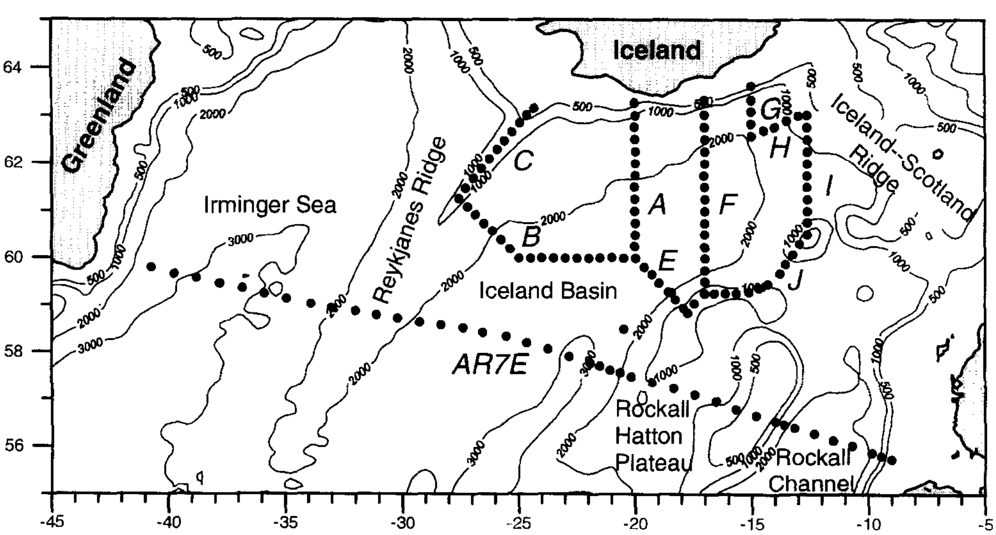

Fig. 1. Position of the hydrographic stations and sections occupied during the DUTCH-WARP cruise of RV "Tyro" in April–May 1991. The depth of the isobaths is indicated in meters.

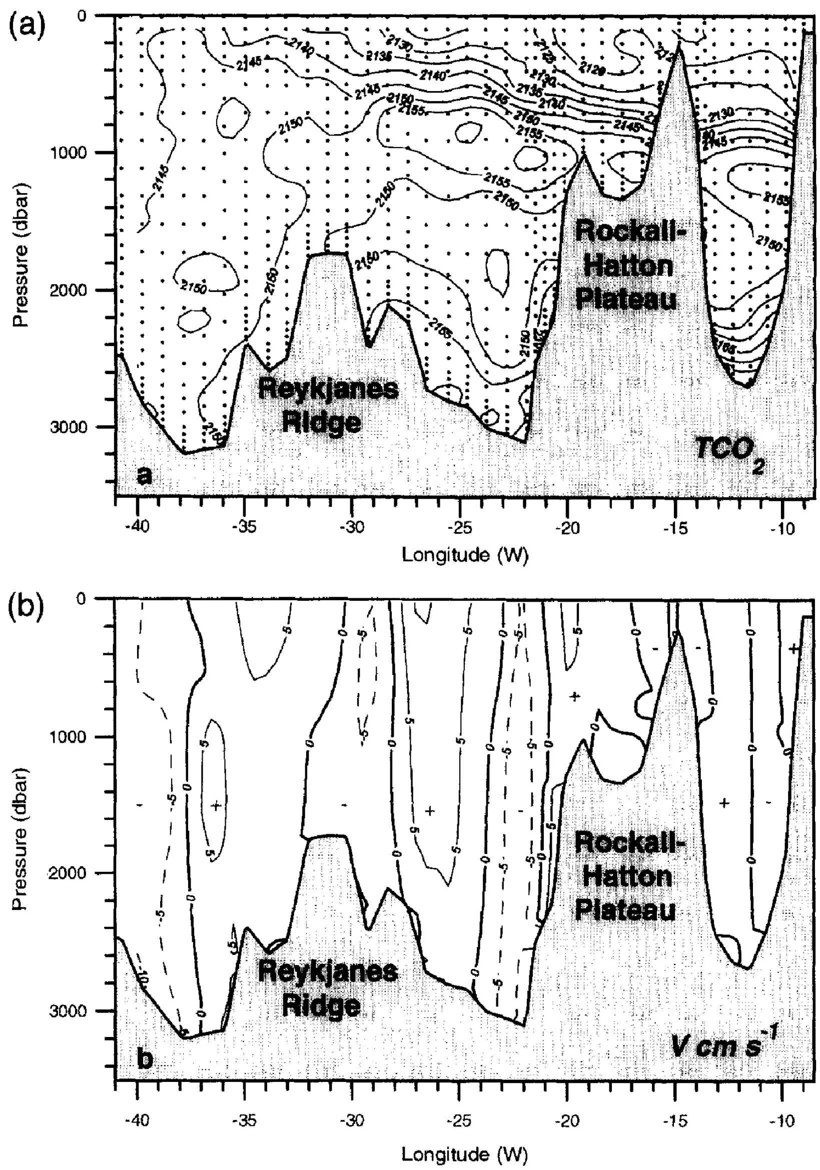

Fig. 2. a. Smoothed distribution of TCO_2 (μmol kg^{-1}) along section AR7E.
b. Mean geostrophic velocity (cm s^{-1}) perpendicular to section AR7E from the Monte Carlo experiment with the extended inverse model.

(1995), while the hydrographic characteristics of the TCO_2 distribution are described by Stoll et al. (1996-this issue).

2. Decomposition of advective fluxes

The geostrophic flow in the ocean across a zonal section can generally be divided into a baroclinic part, connected with the along-section density gradients and a barotropic part due to the slope of the sea surface. At their turn, each of these components can be divided again into a section-wide average value and an eddy term. We will use this discrimination to make a type of Reynolds decomposition of the advective fluxes, which is an extension of the method used by Hall and Bryden (1982) for the determination of the meridional heat flux.

We assume that the surface of the section is S, that x is the distance coordinate along the section and that z is the vertical coordinate. Let V be the cross-sectional velocity and C the volumetric concentration of a substance (e.g., TCO_2). The mean barotropic velocity $\langle V \rangle$ is defined as:

$$\langle V \rangle = \frac{1}{S}\int_S V \mathrm{d}S \tag{1}$$

The spatially variable part of the velocity, defined as $V_v = V - \langle V \rangle$, consists of the depth-dependent mean baroclinic velocity, responsible for the basin wide overturning, $V_{ov}(z)$, the baroclinic eddy velocity, $V_{bce}(x)$, and the barotropic eddy velocity, $V_{bte}(x,z)$. $V_{ov}(z)$ is defined as:

$$V_{ov}(z) = \frac{1}{(x_2 - x_1)}\int_{x_1(z)}^{x_2(z)} V_v(x,z)\mathrm{d}x \tag{2}$$

where $x_1(z)$ and $x_2(z)$ are the boundaries of the section at a vertical level z. Note that V_{ov} can be decomposed further into a geostrophic contribution, V_{ovg}, and a contribution from the directly wind-driven Ekman transport in the surface layer, V_{Ek}. The barotropic eddy velocity $V_{bte}(x)$ is defined as:

$$V_{bte}(x) = \frac{1}{h(x)}\int_{-h(x)}^{0} (V_v(x,z) - V_{ov}(z))\mathrm{d}z \tag{3}$$

where $h(x)$ is the water depth at position x. The remaining baroclinic eddy velocity is defined as:

$$V_{bce}(x,z) = V_v(x,z) - V_{ov}(z) - V_{bte}(x) \tag{4}$$

The concentration $C(x,z)$ is composed similarly of a section-wide average $\langle C \rangle$, an only depth-dependent part $C_{ov}(z)$, an only x-dependent part $C_{bte}(x)$, and a residual $C_{bce}(x,z)$. With this decomposition it is easy to show that the total advective transport of the substance C across the section with surface S equals:

$$\begin{aligned} \mathrm{Tr}_C &= S \cdot \langle V \rangle \cdot \langle C \rangle + \int_S \{V_{ovg} \cdot C_{ov} + V_{Ek} \cdot C_{ov} \\ &\quad + V_{bce} \cdot C_{bce} + V_{bte} \cdot C_{bte}\}\mathrm{d}S \\ &= \mathrm{Tr}_{net} + \mathrm{Tr}_{ovg} + \mathrm{Tr}_{Ek} + \mathrm{Tr}_{bce} + \mathrm{Tr}_{bte} \end{aligned} \tag{5}$$

The total transport therefore can be seen as a combination of a transport, caused by the mean net flow across the section, Tr_{net}, a transport caused by basin-wide baroclinic overturning, Tr_{ovg}, and by the wind-driven Ekman flow, Tr_{Ek} as well as transports due to laterally correlated variations of the barotropic and baroclinic flow with C, Tr_{bce} and Tr_{bte}.

3. Determination of the flow field

For the computation of the geostrophic flow across section AR7E we have used different methods: computation relative to a fixed level of no motion, as well as two inverse models. The first inverse model has quite simple mass-balance constraints while the second model has a whole diverse suite of constraints. With the inverse models a reference velocity in the reference level has been estimated for each station pair, written as the sum of a series expansion obtained from the singular value decomposition of the constraints (Wunsch, 1978; Fiadeiro and Veronis, 1982).

Firstly, we have calculated the geostrophic flow relative to a level of no motion, fixed to a potential density surface. For the choice of such a level we used information we obtained from the Iceland Basin. With an Empirical Search Method (Fiadeiro and Veronis, 1982; Sy, 1988) applied to eight density layers in the boxes, set up by the hydrographic sections in the Iceland Basin (Fig. 1), we found that the density level with $\sigma_\Theta = 27.725$ kg m^{-3} as level of no motion minimized the mass flux imbalance for the boxes in the Iceland Basin. This agrees with the findings by Van Aken (1995) that this reference level gave an optimal agreement between geostrophic velocity and mean velocity as determined with current meters.

Using the level of no motion at $\sigma_\Theta = 27.725$ kg m^{-3} the net northward flow across AR7E amounted to 2.1 Sv. The narrow East Greenland Current, not covered by our survey of AR7E, transports about -4 Sv southwards (Hopkins, 1991). The wind-driven mean Ekman transport between Greenland and Ireland is also directed southwards, given the prevailing western winds over the northern North Atlantic. From the zonally averaged wind stress for the North Atlantic, published by Isemer and Hasse (1987), we have determined the annual mean wind stress between 55°N and 60°N to be 0.08 ± 0.02 N m^{-2}. Over the distance between Ireland and Greenland (~ 2100 km) this will result in a mean southward

Ekman transport of 1.4 ± 0.35 Sv. Assuming a zero net flow between Ireland and Greenland, the net northward geostrophic flow across section AR7E should be 3.3 Sv higher than obtained with a level of no motion. Assuming a realistic additional net flow because of the net transport through Bering Strait (0.8 Sv; Coachman and Aagaard, 1988) only had an effect on the barotropic terms Tr_{net} and Tr_{bte}, but the latter effect was negligible compared to the other transport terms. Therefore we used the zero net flow constraint in a simple inverse model like the models used by Roemmich and Wunsch (1985) and Klein et al. (1995), together with the constraints that in the Iceland Basin the southward flow below the $\sigma_\Theta = 27.8$ kg m^{-3} surface amounts to 3.5 Sv (Saunders, 1993; Dickson and Brown, 1994; Van Aken, 1995) and that in the Irminger Sea the southward transport below this density level amounts to 12 Sv (Dickson and Brown, 1994). This simple inverse model was used for a second calculation of the geostrophic velocity. The resulting reference velocities in the $\sigma_\Theta = 27.725$ kg m^{-3} surface had a RMS value of 1 cm s^{-1}.

Such simple constraints as used for our second calculation only permit estimates of the absolute velocities on large spatial scales (Roemmich and Wunsch, 1985). We introduced additional constraints in an extended inverse model for a third calculation of the geostrophic flow with more detail on basin scales (e.g., barotropic recirculation within basins). The Lower Deep Water (LDW), characterized by high dissolved silica values due to its large Antarctic Bottom Water content, as well as the low-salinity Labrador Sea Water (LSW) are assumed not to participate in the exchange between the Atlantic Ocean and the Nordic Seas (McCartney, 1992; Van Aken, 1995; Stoll et al., 1996-this issue). Therefore we added as extra constraints no net flow of water within the high dissolved silica core (Si > 12 μmol kg^{-1}) originating from the LDW in the Irminger Sea, the Iceland Basin and the Rockall Channel separately; and no net meridional flow in the low-salinity cores of the LSW in the three basins separately ($S < 35.00$ in the Rockall Channel, $S < 34.95$ in the Iceland Basin, and $S < 34.90$ in the Irminger Sea). Further we used the fact that because of the conservation of potential vorticity the barotropic flow on an f-plane will follow the depth contours. This implies that the barotropic transport in the deep basins will not pass the shallow sills (500–850 m) in Denmark Strait and on the Iceland–Scotland Ridge. Also the lateral barotropic exchange between the different basins is limited by the depth of the Reykjanes Ridge and the Rockall Hatton Plateau north of section AR7E. Because the AR7E section runs nearly zonally this effect was formulated as constraints in the inverse model, requiring zero barotropic transport between identical bottom depth contours, from 1500 m downward in 500-m steps.

Inverse models may be quite sensitive for slight variations in the mean depth between hydrographic station pairs, used in the constraints (Stommel and Veronis, 1981). To investigate the effects of this sensitivity for the total carbon flux we have performed a Monte Carlo experiment in which we added a Gaussian noise term with a standard deviation of $\pm 5\%$ to the depths used in the extended inverse model. With the typical water depths along section AR7E this is equivalent with errors in the water depth of the order of 100–150 m, which is a worst-case scenario for the determination of the mean depth between two stations. From this Monte Carlo simulation the full solution of this inverse model with 16 equations and 39 unknown reference velocities appeared to be rather sensitive for small changes in the constraints. Therefore we truncated the solution at 14 singular values, which reduced the typical uncertainty in the reference velocity between two stations with a $\pm 5\%$ noise in the constraints from 4.2 cm s^{-1} for the full solution to 1.8 cm s^{-1} for the truncated solution. The resulting reference velocities (Fig. 2b) had a RMS value of 5.9 cm s^{-1}, a value of the same order of magnitude as the eddy velocity of transient barotropic eddies in the Iceland Basin, measured with current meters (Van Aken, 1995).

The characteristic inflow of LDW with velocities of > 5 cm s^{-1} over the western slope of the Rockall–Hatton Plateau, resulting from the extended inverse model, was confirmed from the hydrography (Van Aken and De Boer, 1995) and current measurements (Van Aken, 1995). Because of the potential vorticity constraints and the salinity and silica constraints, the extended inverse model showed a barotropic circulation in the Irminger Sea with a cyclonic character (Fig. 2b), and with a magnitude of 21.3 Sv. This value is in fair agreement with the barotropic, cyclonic, circulation of 23.1 Sv, determined by Bersch and Meincke (1994) and Bersch (1995) from a nearly identical hydrographic section in the Irminger Sea. Instead of an inverse model, they used ship-borne ADCP measurements to determine the reference velocities. The flow, resulting from the limited inverse model, strongly underestimated this cyclonic barotropic circulation at 7.7 Sv.

4. Components of meridional carbon transports

Following Brewer et al. (1989), Broecker and Peng (1992), and Keeling and Peng (1995), we assume here that the net meridional volume transport between Greenland and Ireland equals zero. Therefore the term $Tr_{net} = S \cdot \langle V \rangle \cdot \langle TCO_2 \rangle$ from Eq. (5) also is assumed to equal zero. The effect of deviations from this assumption for the carbon budget of the North Atlantic is discussed in Section 6. The remaining transport terms for section AR7E have been calculated according to Eq. (5). This leaves out the contribution of the East Greenland Current to the meridional càrbon transport. Since our section AR7E ended just short of that current we will use the values of TTO stations 183 to 187 in the East Greenland Current, observed in August 1981, to determine the contribution of the East Greenland Current. In general the TTO stations in this region show good agreement with our values (Stoll et al., 1996-this issue). The mean value of TCO_2 for the samples from the upper 230 m of these stations with a positive apparent oxygen utilisation, that is, not directly influenced by primary production as in the upper 80 m, was 2133 ± 4 μmol kg^{-1}. The $\langle TCO_2 \rangle$ for our section AR7E was 2144 μmol kg^{-1}. This leads to a typical deviation in the East Greenland Current of $C_v = -11 \pm 4$ μmol kg^{-1} when excluding an expected increase of TCO2 between 1981 and 1991. The southward (= negative) volume transport of the East Greenland Current is assumed to be −4 Sv. With the value of C_v in the East Greenland Current this leads to a northward (= positive) transport of dissolved inorganic carbon of $Tr_{EGC} = 0.05 \cdot 10^6 \pm 0.02 \cdot 10^6$ mol s^{-1}.

The transport terms Tr_{ovg}, Tr_{Ek}, Tr_{bce}, and Tr_{bte} according to the different estimates of the reference velocity have been computed (Table 1). The transport, due to the net overturning as well as the baroclinic eddies, ($Tr_{ovg} + Tr_{Ek}$, and Tr_{bce}), appear to depend only in a limited way from the method by which the velocity has been estimated. Tr_{ovg} mainly depends on the geostrophic shear and therefore on the observed density field, and only for a minor part on the constraints, used in the inverse model. The Ekman contribution to the overturning inorganic car-

Table 1
Estimates of the meridional carbon transport terms at 58°N defined in Eq. 5 in 10^6 mol s^{-1} — upper part (a) for total carbon dioxide and lower part (b) for dissolved organic carbon, assuming the difference between surface and deep water to be either 10 or 40 μmol kg^{-1}

(a) Total carbon dioxide

Velocity estimate	Tr_{ovg}	Tr_{Ek}	Tr_{bce}	Tr_{bte}	Tr_{EGC}	Tr_C
Relative $\sigma_\Theta = 27.725$ kg m^{-3}	−0.28	+0.03	+0.03	−0.01	+0.05	−0.18
Limited inverse model	−0.28	+0.03	+0.02	0.00	+0.05	−0.18
Extended inverse model	−0.27	+0.03	+0.03	+0.09	+0.05	−0.07
Monte Carlo experiment (±5%)	−0.27	+0.03	+0.03	+0.10	+0.05	−0.06
Uncertainty	±0.01	±0.01	±0.003	±0.05	±0.03	±0.06

(b) Dissolved organic carbon estimated with the extended inverse model

Difference between surface and deep water	Tr_{ovg}	Tr_{Ek}	Tr_{bce}	Tr_{bte}	Tr_{EGC}	Tr_C
ΔDOC = 10 μmol kg^{-1}	+0.10	−0.03	–	–	−0.03	+0.04
to	to	to			to	to
ΔDOC = 40 μmol kg^{-1}	+0.40	−0.12			−0.12	+0.16

Positive values are northward, negative values are southward transport.

bon transport is directed to the north (positive) with a magnitude of only ~10% of the southward directed geostrophic overturning. While, as can be expected, the overturning supports a southward carbon transport, the baroclinic eddy term is, although small, in a northward direction. As can be expected the barotropic contribution to the total transport, Tr_{bte}, depends strongly on the method to determine the reference velocity. Both with a level of no motion as with the limited inverse model the barotropic recirculation in the basins is under-estimated. The realistic cyclonic barotropic circulation in the Irminger Sea, obtained with the extended inverse model, correlates well with the existing vertically averaged TCO_2 gradient (Fig. 2a), supporting ~80% of the barotropic contribution to the total TCO_2 transport, Tr_{bte}.

With the Monte Carlo simulation, described above, we also assessed the sensitivity of the TCO_2 transport for a ±5% noise in the constraints. The resulting mean values of Tr_{ovg} and Tr_{bce} as well as Tr_{bte} hardly differ from the direct results of the extended inverse model. The uncertainty of the TCO_2 transport terms expressed as the standard deviation from the Monte Carlo experiment [Table 1, (a)] indicate that the main uncertainty in the estimates is in the barotropic contribution. We propose here the result of the Monte Carlo experiment, that is a small southward transport of dissolved inorganic carbon of $-0.06 \cdot 10^6$ mol s^{-1} (0.02 Gt a^{-1} C) as the best possible estimate. The southward overturning term, $Tr_{ovg} + Tr_{Ek}$ is $-0.24 \cdot 10^6$ mol s^{-1}, a factor 1.5 higher than the $-0.16 \cdot 10^6$ mol s^{-1}, estimated by Stoll et al. (1996-this issue) from a rather crude budget of the overturning. However, the contributions from the barotropic circulation and the East Greenland Current compensate the overturning transport for a large part, resulting in a quite low value for Tr_C. We assume that the ±5% uncertainty in the constraints used in the Monte Carlo simulation is realistic. Also, given the uncertainty in the contributions of the Ekman transport and the properties of the East Greenland Current, the resulting uncertainty in the estimate of Tr_C is of the same magnitude as the estimated value of Tr_C [Table 1, (a)]. Although a rigorous statistical analysis is not possible with these uncertainty estimates, these results at least indicate that Tr_C is essentially small and possibly does not significantly differ from zero. This is mainly due to the uncertainty in the laterally varying barotropic flow, and therefore to the uncertainty of Tr_{bte}.

5. Transport of dissolved organic carbon

Next to the TCO_2 in the ocean there exists a dissolved organic carbon pool which may contribute to the total advective carbon flux. The DOC concentration is known to be higher in the surface layers of the ocean, compared to the deep ocean (Duursma, 1963; Kepkay and Wells, 1992; De Baar et al., 1993; Carlson et al., 1994). As with the TCO_2, such a spatially varying DOC distribution may contribute to the total advective transport of total dissolved carbon, probably northwards due to the large-scale overturning in the North Atlantic Ocean.

In order to asses the effect of the overturning transport of DOC for the total dissolved carbon transport we want to estimate the order of magnitude of the DOC contribution. Within the north Atlantic Ocean the difference of DOC between the surface layers and the deep ocean may vary from 10 μmol kg^{-1} along a section off Cape Farewell (Duursma, 1961), 15 μmol kg^{-1} in the Sargasso Sea (Carlson et al., 1994) to 40 μmol kg^{-1} in the Iceland Basin (De Baar et al., 1993). Duursma (1961) reported an April 1958 section with uniform DOC values at ~60 μmol kg^{-1} and a September 1958 section where surface waters exhibited elevated DOC levels as high as ~100 μmol kg^{-1}. This is consistent with seasonal increases of surface water DOC following plankton blooms as reported elsewhere (Duursma, 1961; Copin-Montégut and Avril, 1993; Carlson et al., 1994). The transition between these layers was found in the Iceland Basin at a depth of 500 m coinciding with the upper side of the permanent pycnocline (De Baar et al., 1993). We applied this distribution to our section AR7E and calculated the DOC transport across this section with the geostrophic velocities from the extended inverse model described above. Here it is noted that the analytical methods for determining DOC in seawater are still being developed and not as accurate as the TCO_2 methods. Briefly, the analytical blank of DOC methods may easily be underestimated. Otherwise it now appears that deep-water values of ~50–60 μmol kg^{-1} and seasonally higher surface water values approaching 70–80 μmol kg^{-1} as recently reported by several laboratories (Copin-Montégut and Avril, 1993; Carlson et al., 1994) are consistent with the original values of Duursma (1961). The absolute DOC values do not affect the result of the calculation however, only the difference between surface layer and the deep ocean. This cancels possible errors due to too high overall blanks. For this difference we applied a value ranging from 10 μmol kg^{-1} (Duursma, 1961; Carlson et al., 1994) to 40 μmol kg^{-1} (De Baar et al., 1993; Duursma, 1961). Our calculations resulted in [Table 1, (b)] a northward DOC transport of $+0.10 \cdot 10^6$ to $+0.40 \cdot 10^6$ mol s^{-1}, because of the large-scale geostrophic overturning, and a southward transport due to the Ekman drift of $-0.12 \cdot 10^6$ to $-0.03 \cdot 10^6$ mol s^{-1}. Also the southward flowing East Greenland Current supports a southward DOC transport of $-0.12 \cdot 10^6$ to $-0.03 \cdot 10^6$ mol s^{-1}. The latter value may overestimate the contribution of the East Greenland Current, because part of the southward flow may be below the rather shallow pycnocline (100 m) in the East Greenland Current. However, this result indicates that for a large part the northward transport of DOC across section AR7E may be of the order of $0.10 \cdot 10^6$ mol s^{-1}. If we combine the DOC transport [Table 1, (b)] with the estimated TCO_2 transport [Table 1, (a)] a very small value results for the transport of dissolved carbon due to the partially compensating effects of the southward TCO_2 transport and the northward DOC transport.

6. Discussion

The estimates of the southward meridional transport of TCO_2 at 58°N between Ireland and Greenland have been obtained with different flow estimates, as described above. The estimates vary between $-0.18 \cdot 10^6$ mol s^{-1} for geostrophic flow estimated with the empirical search method and the limited inverse model and $-0.06 \cdot 10^6$ mol s^{-1} from the Monte Carlo experiment with the extended inverse model. The circulation across the AR7E section, computed with the extended inverse model, has a number of properties in agreement with knowledge from independent sources, for example, northward flow of LDW over the slope of the Rockall Hatton Plateau, and a cyclonic circulation of the order of 22 Sv in the Irminger Sea. The most reliable value for the TCO_2 transport therefore is assumed to be $-0.06 \cdot 10^6$ mol s^{-1}, obtained with the Monte Carlo simulation of the extended inverse model. This low transport value results from compensation of the large-scale overturning transport by transport in the East Greenland Current as well as the barotropic transport, mainly in the Irminger Sea. In the calculations by Brewer et al. (1989) part of the barotropic contribution, due to the anti-cyclonic circulation in the sub-tropical gyre, was covered by the explicit discrimination between transport through the Florida Straits and through the mid-ocean section at 25°N. Broecker and Peng (1992) as well as Keeling and Peng (1995) only allowed for the overturning transport and ignored any contribution from the laterally varying barotropic flow. The main uncertainty in our transport estimate is due to the uncertainty of this zonally varying barotropic flow, and is of the order of magnitude of the net TCO_2 transport. We may conclude that anyhow the southwards transport of TCO_2 across 58°N is between a number of the order of $-0.1 \cdot 10^6$ mol s^{-1} and zero, already largely due to the biological pump (Stoll et al., 1996-this issue). This number is quite small compared with the uptake of CO_2 from the atmosphere in the temperate North Atlantic, estimated by Tans et al. (1990) to be of the order of $0.8 \cdot 10^6$ mol s^{-1}. The physical pump of the oceanic conveyer belt is much less effective for the meridional transport of carbon in the ocean from the cold surface waters at high latitudes to the deep ocean at lower latitudes than the biological pump in the warmer temperate Atlantic Ocean.

The rough estimate of the transport of DOC across section AR7E shows that this transport may be of the same order of magnitude as the net transport of inorganic carbon, but directed to the North. This effect is due to the covariance of the large-scale overturning motion with the assumed vertical stratification of DOC. More recently measured profiles in the Pacific Ocean exhibit a similar trend of elevated DOC in the surface water, with concentrations of 70 μmol kg^{-1} in surface waters and 40 μmol kg^{-1} in deep waters (Sharp et al., 1995). When applied to our study, the here assumed similar magnitude of the vertical stratification of DOC between 10 and 40 μmol kg^{-1} yields a significant impact on the assessment of overall meridional carbon transport. Obviously the current drive towards more accurate DOC values in seawater is crucial for proper assessment of carbon transports in the North Atlantic and other oceans.

An implicit assumption of the calculations presented by us as well as by Brewer et al. (1989), Broecker and Peng (1992), and Keeling and Peng (1995) is the net zero flow assumption, that is, the assumption that $\langle V \rangle$ equals zero between the eastern and western sides of the Atlantic Ocean. However, Coachman and Aagaard (1988) have proposed a mean volume transport of -0.8 Sv flows through Bering Strait from the Pacific to the Atlantic Ocean, while in the Polar and Arctic Seas ~ 5.400 km^3 $a^{-1} = -0.17$ Sv is added due to river runoff and precipitation excess (Aagaard and Carmack, 1989). If all this water passes between Ireland and Greenland as a southward mean flow, that is $S \cdot \langle V \rangle \approx -1$ Sv, this will lead to an extra southward transport of dissolved inorganic carbon $\mathrm{Tr}_{net} \approx -2.15 \cdot 10^6$ mol s^{-1} (-0.83 Gt a^{-1} C). Estimates of TCO_2 transports as presented by Brewer et al. (1989), Broecker and Peng (1992), Keeling and Peng (1995) and by us (Table 1) ignore the contribution from Tr_{net}. The freshwater input from river runoff and precipitation excess in the Polar and Arctic seas, indicative for the divergence of the freshwater transport in the ocean, still will support a southward inorganic carbon transport of the order of $\mathrm{Tr}_{net} = -0.37 \cdot 10^6$ mol s^{-1} (-0.14 Gt a^{-1} C).

6.1. Towards a North Atlantic carbon budget?

One can argue whether the transport estimates, given by us, combined with the estimate by Brewer et al. (1989) can be used for a reliable estimate of the net carbon budget for the temperate North Atlantic Ocean between 25°N and 58°N. In order to asses this possibility and the uncertainties in such an estimate we will introduce here a simple budget model.

We will consider the ocean as a meridional channel, and assume the meridional transport of carbon in this ocean to be composed of the term Tr_{net}, connected with the net meridional volume flow $S \cdot \langle V \rangle$, and an effectively diffusive term, Tr_{diff}, representing the transport processes without net meridional volume flow. This is written, according to Eq. (5), as:

$$\begin{aligned}\mathrm{Tr}_C &= \mathrm{Tr}_{net} + \mathrm{Tr}_{ovg} + \mathrm{Tr}_{Ek} + \mathrm{Tr}_{bce} + \mathrm{Tr}_{bte} \\ &= S \cdot \langle V \rangle \cdot \langle C \rangle + \mathrm{Tr}_{diff} \qquad (6)\end{aligned}$$

The divergence of this transport is connected with the flux density per latitude unit of the transport of CO_2 into the ocean due to air–sea gas exchange, F_{as}:

$$\begin{aligned}\frac{d(\mathrm{Tr}_C)}{dy} &= \langle C \rangle \frac{d(S \cdot \langle V \rangle)}{dy} + S \cdot \langle V \rangle \frac{d\langle C \rangle}{dy} + \frac{d\mathrm{Tr}_{diff}}{dy} \\ &= F_{as} \qquad (7)\end{aligned}$$

where y is the meridional coordinate. Transport of DOC, burial into the sediment as well fluxes from the sediment have been ignored here as well as other source and sink terms like the input into the oceans of dissolved carbon by river input and precipitation. From Eq. (7) we see that the divergence of the meridional carbon flux is connected with the total water budget via the divergence of the net meridional volume transport, written as:

$$\frac{d(S \cdot \langle V \rangle)}{dy} = (P - E) + R \qquad (8)$$

where P and E are the precipitation and evaporation density, respectively, and R is the flux density of fresh water due to river runoff. Integration of Eqs. (7) and (8) between two latitudes with coordinates y_1 and y_2 where estimates of Tr_{diff} are available results in:

$$\underset{(A)}{\int_{y_1}^{y_2} F_{as} dy} = \underset{(B)}{\int_{y_1}^{y_2} \langle C \rangle [(P-E)+R] dy} + \underset{(C)}{\int_{y_1}^{y_2} S \cdot \langle V \rangle \cdot \frac{d\langle C \rangle}{dy} dy} + \underset{(D)}{\mathrm{Tr}_{diff}(y_2)} - \underset{(E)}{\mathrm{Tr}_{diff}(y_1)} \qquad (9)$$

The order of magnitude of the different terms can be estimated from the literature. According to Tans et al. (1990), the term on the left-hand side (A) is of the order of $A = 0.78 \cdot 10^6$ mol s^{-1} (0.30 Gt a^{-1} C) for the temperate North Atlantic Ocean between 15°N and 50°N. We assume here that the order of magnitude of this value also can be applied to the northward shifted area between 25°N and 58°N. The first integral on the right-hand side (B) can be estimated from precipitation, evaporation and runoff data. The net evaporation minus precipitation in the North Atlantic Ocean between 25°N and 58°N amounts to 0.35 Sv while the river runoff to the Atlantic in this latitude belt equals 0.12 Sv (Baumgartner and Reichel, 1975). With a typical $\langle TCO_2 \rangle$ value of 2150 μmol kg^{-1}, this gives a value of about $B = -0.51 \cdot 10^6$ mol s^{-1} (-0.20 Gt a^{-1} C). With a typical net southward meridional transport of the order of 1 Sv, as presented above, we can make an estimate of the second integral (C), given a $\langle TCO_2 \rangle$ value of 2144 μmol kg^{-1} at 58°N from our data, and at 25°N of ~ 2168 μmol kg^{-1} from Brewer et al. (1989). From these values a low value of $C = +0.03 \cdot 10^6$ mol s^{-1} ($+0.01$ Gt a^{-1} C) is found. For the transport of inorganic carbon we will use $Tr_{diff}(y_1 = 25°N) = -0.68 \cdot 10^6$ mol s^{-1} (-0.26 Gt a^{-1} C; Brewer et al., 1989), while our results (Table 1) give $Tr_{diff}(y_2 = 58°N) = -0.06 \cdot 10^6$ mol s^{-1} (-0.02 Gt a^{-1} C).

The simple budget model, presented above, clearly shows some striking aspects of the oceanic carbon budget. Firstly we note the large order of magnitude of term B from Eq. (9). This shows that, because of the high background concentration of TCO_2, the divergence of the net meridional volume transport, due to the freshwater balance of the ocean, may generate terms in the oceanic carbon budget that clearly cannot be ignored. Therefore it is not allowed simply to subtract the meridional transports based on a zero net flow assumption at two different latitudes to obtain an estimate of the net oceanic CO_2 uptake. The effects of the freshwater flux have to be accounted for.

Secondly, as shown in Section 5, the net meridional flux of DOC may be of the same order of magnitude as the net flux of TCO_2. This also applies to the divergence of the fluxes, indicating that in order to determine the carbon budget of the ocean from meridional flux estimates, accurate estimates of the meridional DOC flux are needed too. This will require DOC determinations on zonal sections with appropriate accuracy and sample density to allow the determination of such a DOC transport.

Thirdly, the sum of the terms B to E amounts to a total of $A = +0.14 \cdot 10^6$ mol s^{-1} ($+0.05$ Gt a^{-1} C). This value, much smaller than the estimate by Tans et al. (1990) of $A = +0.78 \cdot 10^6$ mol s^{-1} ($+0.30$ Gt a^{-1} C), seems to suggest that the net CO_2 uptake of the North Atlantic Ocean hardly plays a role in the global CO_2 budget. But given the unknown contribution of the DOC transport and the as yet limited accuracy of estimates of the evaporation, and certainly the precipitation over the oceans, the discrepancy between our net carbon budget of the temperate North Atlantic, and that of Tans et al. (1990), mainly stresses that still much high-quality data have to be gathered before a reliable estimate of the oceanic carbon budget can be made. Moreover, given the uncertainty of the inverse model results of the barotropic eddy term Tr_{bte}, as indicated in Table 1, (a), also the uncertainty in the general circulation should be reduced, as stressed by Martel and Wunsch (1993).

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